Dating shallow thrusts with zircon (U-Th)/He thermochronometry—the shear heating connection

Matteo Maino[1], Leonardo Casini[2], Andrea Ceriani[1, 3], Alessandro Decarlis[4], Andrea Di Giulio[1], Silvio Seno[1], Massimo Setti[1], and Finlay M. Stuart[5]

[1]Dipartimento di Scienze della Terra e dell'Ambiente, Università di Pavia, via Ferrata 1, 27100 Pavia, Italy

[2]Università di Sassari, DiSBEG, via Piandanna, 4, 07100 Sassari, Italy

[3]The Petroleum Institute, P.O. Box 2533, Abu Dhabi, United Arab Emirates

[4]IPGS/EOST, rue Blessig 1, F-67084 Strasbourg Cedex, France

[5]Isotope Geosciences Unit, SUERC, Scottish Enterprise and Technology Park, Rankine Avenue, East Kilbride G75 0QF, UK

ABSTRACT

New zircon (U–Th)/He (ZHe) ages from a shallow (<6-7 km) thrust fault zone and surrounding wall rocks in the Helminthoid Flysch of the Ligurian Alps were measured to test the applicability of the thermochronometer for dating the brittle or brittle-ductile faults. The ages are integrated with X-ray diffraction analysis of clay minerals and fluid inclusion microthermometry on vein-filling minerals to constrain the temperature conditions of the damage zone and the wall rocks during thrusting.

The wall rocks yield pre-depositional inherited ZHe ages (125.3±15 to 312.3±37 Ma) while ages from the fault core are reset (28.8±3.4 to 33.8±4.0 Ma). This is consistent with independent geological and thermochronometric evidence for Early Oligocene motion of the thrust. This implies that the fault zone exceeded 200°C during faulting, and confirms the illite

crystallinity and fluid inclusions constraints, which indicate temperatures of 220-300°C in the fault zone, while in the wall rocks were <180-200°C.

Thermal modeling of the fault zone suggests that the shear heating associated with the fault motion is an efficient mechanism for generating temperature increases of 50-70° during a displacement of 10-25 km in 2–10 Ma. Our results underscore the validity of ZHe technique for dating brittle or brittle-ductile faults characterized by relatively high strain rate.

INTRODUCTION

Determining the timing of brittle and brittle-ductile fault activity is essential for reconstructing tectonic processes. A number of studies have focused on the potential and the application of thermochronometric techniques (^{40}Ar-^{39}Ar, K-Ar, Rb-Sr) to determine the absolute timing of specific thermal stages experienced by faults during their evolution (e.g. Lyons and Snellenberg, 1971; van der Pluijm et al., 2001; Torgersen et al., 2014). These techniques capitalize on authigenic, syn-kinematic clay minerals found in brittle cataclasites. Until now, ^{40}Ar-^{39}Ar and K-Ar dating of authigenic illite have proved to be the best methods for determining the timing of near-surface deformation and brittle or brittle-ductile faulting at low temperatures (van der Pluijm et al., 2001; Clauer et al., 2012). The technique records rock cooling through 100-300°C, depending on clay minerals and polytypes (e.g. Haines and van der Pluijm, 2008; Torgersen et al., 2014). Although it has been successfully applied in various geological contexts (e.g. Zwingmann et al., 2010a, b; Viola et al., 2013) several problems may afflict the method. For instance, Ar can be lost by thermal diffusion or by exchange with hydrothermal fluids (Torgersen et al., 2014; 2015). Additionally, the clay-rich gouges may contain detrital clays that must be removed from authigenic illite in order to avoid measuring mixed ages (Clauer and Chaudhuri, 1995; van der Pluijm et al., 2001).

Absolute dating of brittle faults with low temperature thermochronometers (apatite and zircon fission track and (U-Th)/He dating) has only been applied to pseudotachylites where high temperature produces frictional melting and leads to complete resetting even in a nearly instantaneous event (e.g. Murakami et al. 2006). Apart from this case, the heat generation produced during localized, short-lived faulting is generally considered insufficient to reset mineral systems (d'Alessio et al., 2003). However, distributed fault damage zones can be characterized by temperature increases of up to 150°C with respect to the wall rocks (Morton et al., 2012). If this occurs on timescales of 10^5-10^6 years, low temperature mineral thermochronometers may be reset in the fault zone, yielding a method for dating fault movement. However, there are several mechanisms - e.g. advection from a warm body, circulation of hot fluids or shear heating - that can generate heating (Morton et al., 2012; Ben-Zion and Sammis, 2013). Quantifying how these mechanisms contribute to heating is a challenge, which may be addressed by a comprehensive thermal characterization throughout the fault zone.

Here we test whether zircon (U-Th)/He (ZHe) dating can be used to determine when a long-lived thrust fault developed in the shallowest part of an orogenic wedge. The ZHe thermochronometer has a well-behaved He partial retention zone (HePRZ) of 130–200°C and a closure temperature (Tc) of ~180°C (Wolfe and Stockli, 2010). Consequently, it is ideally suited to dating large heat-producing faults that were active at shallow depths (<6-7 km) where wall-rock temperature does not exceed Tc.

This study combines illite crystallinity data and fluid inclusion analyses of vein-filling minerals to estimate the temperatures experienced by the fault zone and wall rocks from the southernmost segment of the Penninic Front in the Ligurian Alps (Fig. 1). ZHe ages are determined from fault zone and wall rock samples, with the aim of testing the suitability of the thermochronometer to recording the time temperature change occurred during faulting. The potential causes of heating - shear heating or circulation of deep hot fluid - and how they are related to fault motion are explored analyzing the thermal signatures resulting from the dataset and verified through a set of numerical models.

GEOLOGICAL SETTING

The fault selected for our study is part of the composite Penninic Front of the Alps (Fig. 1A). This complex fault-system represents the contact between the external and internal zones of the Alpine belt (e.g. Ceriani et al. 2001). In the study area (i.e. Ligurian Alps; Fig. 1B-C) the fault system comprises the basal thrust of the Helminthoid Flysch nappes (HF), which are detached oceanic cover units corresponding to the frontal part of the Alpine accretionary wedge (Di Giulio, 1992). During the Eocene, the HF were gravitationally transported NW from the oceanic realm towards more proximal portions of the European foreland. In the Early Oligocene the HF nappes were thrust SW over both the internal Alpine units (Briançonnais) and the foreland basin succession deposited onto the external European crust (Dumont et al., 2012). Ford et al. (1999) estimate that at least 50 km of displacement occurred during the Oligocene but this value may encompass part of the earlier gravitational translation. Thus 50 km can be considered an upper limit on the displacement of the studied thrust.

The Early Oligocene HF emplacement in the Western Alps is constrained by the biostratigraphic and tectono-sedimentary record (e.g. Ford et al., 1999; Ceriani et al., 2001; Simon-Labric et al., 2009). Later deformation phases produced rotation and shallow extensional faulting of the Penninic front (Maino et al., 2012; 2013).

The hanging wall of the study area (Sanremo unit; SRU) is dominantly Cretaceous arkose (Bordighera Sandstone, BS) overlain by calcareous turbidites (San Remo Flysch; Fig. 1C). The pre-erosional thickness of SRU is estimated to be 4-6 km on the basis of thermometric and petrologic data (Maino et al., 2012). The footwall consists of Eocene to earliest Oligocene foreland turbidite deposits of shales and quartz- and mica-rich sandstones (Ventimiglia Flysch, VF). The occurrence at the top of the VF of olistostromes preceding the tectonic emplacement of the SRU provides a biostratigraphically-calibrated constraint on the beginning of the SRU emplacement (earliest Oligocene; Decarlis et al., 2014).

FAULT ARCHITECTURE

The contact between VF and SRU is formed from two NE-dipping faults comprising a chaotic deposit known as Schistes à Blocs (Kerckhove, 1969; Fig. 1C). It consists of a clay-rich tectonosedimentary mélange (Flysch Noir, FN) including up to km-size tectonic sheets made of portions of the Briançonnais succession (Decarlis et al., 2013).

The footwall rocks (VF) are relatively undeformed, affected only by a distribute fracture network and long-wave-length open folds. The SRU rocks of the hanging wall are involved in SW-verging tight to sub-isoclinal folds associated with minor thrusts (Fig. 2D). The fault zone structure is strongly asymmetric (Fig. 2A-D), consisting of a 40-150 m wide damage zone encompassing at its top a ~50 cm thick fault core made of pervasively foliated (scaly fabric) gouges and proto-cataclasites (Woodcock and Mort, 2008). The thin (0.5-2 m) hanging wall damage zone is an extensively fractured and veined cataclastic breccia of BS sandstones characterized by a spaced foliation. FN shales and sandstones of the footwall damage zone are characterized by alternation of foliated gouges, cataclastic breccias and undeformed strata (Fig. 2A-B). The entire damage zone is affected by variably oriented fractures and abundant quartz-calcite veins, which constitute up 5-15% of the outcrop volume. Asymmetric folds and S-C bands show consistent top-to-the SW kinematic (Fig. 2B-C).

ZHE THERMOCHRONOMETRY

(U-Th)/He age determinations[1] were performed on zircons from eleven samples of BS, FN and VF sandstones collected across the fault zone and into the wall rocks (Fig. 1C, 2A, F; Table DR1). ZHe ages of samples from within the fault core (samples T1-4) are between 28.8 ± 3.4 and 33.8 ± 4.0 Ma. Samples from the footwall (F1-2; 307.8 ± 37 and 312 ± 18.9), hanging wall (H2-3; 125.3 ± 15 and 158.7 ± 18.9) and damage zone (H1 and D1-2; between 93.5 ± 11 and 118.7 ± 14 Ma) are considerably older. Only the samples from the fault core have post-depositional ages, whereas hanging wall, footwall and damage zone samples have pre-depositional ages. Helium diffusion modeling[1] indicates that relatively short periods of heating (2-8 Ma) are sufficient to reproduce the measured ages (Fig. DR1).

TEMPERATURE CONDITIONS

XRD Data

Eight clay-rich layers from both the footwall and hanging wall rocks and six from the fault damage zone were subjected to XRD analysis[1] (Fig. 1C-2E; Table DR2). The <2 μm fraction of the wall rocks has an illite crystallinity (0.31-0.45 KI, in agreement with data reported by Piana et al., 2014) that fall in the late diagenetic/low anchizone conditions (150-250°C, Merriman and Frey, 1999) progressively increasing toward the fault zone (Fig. 2E). Within the fault zone the illite crystallinity ranges between 0.26 and 0.31 KI, with the lowest values detected into the fault core and secondary intensively deformed bands. These data suggest anchizone conditions (~250°C) approaching the epizone (~300°C) in correspondence of the highly deformed bands.

Fluid Inclusion Data

Microthermometric measurements[1] have been performed to determine the temperature of fluids flowing during thrusting (Barker and Goldstein, 1990). Homogenization temperatures (Th) were measured on aqueous two-phase inclusions trapped in calcite crystals filling sigmoidal veins from close to the fault core (Fi1) or in the hanging wall (Fi2; Fig. 1C; 2A; Table DR3). Sample Fi1 yields a cluster of values in the range 220 to 270°C and a few isolated lower values down to 137°C (Fig. DR2). Sample Fi2 shows a larger temperature range (130-260°C) with peaks around 130-140°C and 180-200°C. The lowermost values in both samples are related to secondary inclusions, most likely formed during a late tectonic cooling stage.

IS THE THERMAL SIGNAL RELATED TO FAULT MOTION?

Illite crystallinity and fluid inclusion data show that fault zone rocks were heated to 220-300°C, while the wall rocks did not exceed 180-200°C. Nevertheless, only samples from close to the fault core were heated sufficiently and/or for long enough to completely reset the ZHe system (Fig. 2F). Zircons from the damage zone and wall rocks show Carboniferous or Jurassic-Cretaceous ages that are compatible with the late-Variscan tectonics (Casini et al., 2015) or the post-rift fission track data (Danišík et al., 2007) reported for the Corsica-Sardinia basement, which is considered as the source of the BS and VF sandstones (Di Giulio, 1992). However, partial resetting of the ZHe ages is likely to have occurred in the wall rocks and, particularly, within the damage zone where the temperature inferred from XRD and fluid inclusion analyses was higher.

The evidence for cool wall rocks grading toward a hotter fault core is consistent with a thrusting-related overheating focused within the fault zone. That this heating is directly related to fault motion is confirmed by the consistency of the ZHe ages (T1-4) with the Early Oligocene age of the Helminthoid Flysch basal thrust activity inferred from the biostratigraphic and tectono-sedimentary record (Ford et al., 1999; Ceriani et al., 2001) and the phengite ^{40}Ar-^{39}Ar dating of a northern metamorphic segment of the Penninic Front (34-27 Ma; Simon-Labric et al., 2009). Another potential cause, such as lateral or vertical advection must be discarded, as it requires a warm block to transfer heat to and along the fault.

In order to quantify the potential contribution of shear heating and flow of hot fluids to generating the ~40-120°C overheating, we explore the thermal behavior of the fault zone for the two cases using numerical models.

THERMAL MODELING

The evolution of the crustal geotherm during faulting has been modeled[1] using a modified version of the finite-difference code GEOTHERM (Casini et al., 2013). For simplicity, we neglect the thermal effect of brittle processes involving pseudotachylite development (Han et al., 2007; Di Toro et al., 2004) and rock pulverization (Ben Zion and Sammis, 2013). Our results, therefore, provide a conservative estimate of shear heating (Burg and Gerya, 2005).

The first set of experiments (Table DR5) simulates deformation at variable strain rates (between 10^{-15} and 10^{-12}), obtained for realistic values of displacement and duration of thrusting (5 to 50 km, 2-14 Myrs). A second set of experiments (Table DR6) simulates the circulation of hot (300°C) fluids in the fault damage zone. Flow rates are constrained from the density of syn-tectonic veins (expressed as ϕ that is the ratio of the thickness-equivalent volume of syn-tectonic veins and host rock) and the ZHe ages.

All experiments were conducted between 50 and 0 Ma to minimize numerical artifacts related to the choice of initial conditions. In all runs (except for the no-shear experiment) the fault core is characterized by a temperature increment above that of footwall rocks (Fig. DR3). The temperature increase is particularly apparent in the experiments with shear heating.

However, a significant (50-70°C) thermal increment can be maintained for periods long enough to reset the ZHe system only if 10–25 km of displacement was accumulated in 2–10 Myrs (Fig. 3A). These results are consistent with the observed cutoff and the ZHe constraints. Lower bound displacements (10–16 km) accomplished in longer periods of deformation (>6 Myrs) do not produce enough heating. Larger displacement and short periods of fault motion cause, instead, unrealistically high temperatures in the fault core (T>350°C). Of course, the advection of hot fluids might have contributed to the thermal budget (e.g. Morton et al., 2012). However, our experiments (Fig. 3B) show that fluids have a minor effect on the thermal structure of the fault. In fact, temperature increases of >40°C can be obtained only for almost instantaneous advection or very large volumes of fluids (ϕ values between 0.2 and 0.6).

CONCLUSIONS

The basal thrust of the Penninic Front provides a powerful empirical test showing excellent agreement of ZHe ages, thermal and geological constraints. Our results demonstrate the reliability and applicability of ZHe thermochronometry for dating the end of heating related to thrusts in the shallow crust (<6-7 km). Thermal data and modeling provide a quantitative estimation of the potential contribution of shear heating and advecting hot fluids to generating the measured overheating (~40-120°C) within the fault zone. The numerical experiments indicate that the temperature within the core of the Penninic Front might have increased up to 50–70°C if 10-25 km of finite displacement were accumulated in 2 to 10 Myrs, which is consistent with the available geological constraints confirming the effectiveness of shear heating on geological timescales. The potential contribution of hot fluids is quantified to be one order of magnitude lower. On the whole, this methodology has a great potential for dating brittle and brittle-ductile faults characterized by relatively large displacements or rapid deformation.

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FIGURE CAPTIONS

Figure 1. A) Location of the study area. B) Tectonic sketch of the Ligurian and SW Alps. Ar: Argentera massif; Dauph: Dauphinois domain; HF: Helminthoid Flysch; PF: Penninic Front; TPB: Tertiary Piedmont Basin. C) Geological map of the basal thrust of the Helminthoid Flysch. The locations of sampling sites for fluid inclusion (Fi), illite cristallinity (Ic) and zircon (U-Th)/He (ZHe) analyses are shown as stars.

Figure 2. A) Photograph of an upper portion of the studied fault zone. BS: Bordighera sandstones; FN; Flysch noir; hw-dz: hanging wall damage zone; fw-dz: footwall damage zone. Starsc circles and squares show the location of some samples used for ZHe dating, illite cristallinity and fluid inclusion microthermometry, respectively. B) Variably spaced foliation affecting the shales and sandstones of the footwall damage zone. The occurrence of syn-kinematic quartz-calcite veins (v) is shown. Asymmetric folds indicate top-to-the SW kinematic. C) Anastomosed foliation developed within the fault core zone. S-C bands are consistent with top-to-the SW thrusting. D) Sketch of the fault zone showing the main deformation fabrics. The zoom on the main fault indicates within the fault core the development of a scaly fabric associated with a transition to ductile creep, whereas in the damage zones the presence of localized cataclasis and discrete foliation suggests predominant brittle faulting. E, F) ZHe ages and illite crystallinity plotted against distance from the fault core. An average 80m thick fault damage zone is assumed.

Figure 3. A) Time vs. displacement plot of the maximum temperature recorded by the fault core considering the effect of the shear heating (all experiments). C) Time vs. veins/host rock volume ratio plot of the maximum temperature recorded by the fault core considering the circulation of hot (300°C) fluids. In both diagrams the field of the realistic conditions fitting the observed

geological (displacement or density of veins) and measured temperature constraints are projected.

[1]GSA Data Repository item 2015xxx. Analytical procedures and data tables of XRD diffraction, fluid inclusion microthermometric and ZHe analyses and the set up, boundary conditions and approximations used in thermal models are available online at www.geosociety.org/pubs/ft2015.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.